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Water balance of a boreal Scots pine forest

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In terrestrial ecosystems, the amount and availability of water is one of the key factors affecting the net primary productivity and other biological processes of the system. At the SMEAR-II station, we have monitored the water balance of two adjacent micro-catchments since 1997. In this study, we report the long-term measurements of precipitation, throughfall, snow depth, soil water content, runoff and evapotranspiration and the annual water balances based on these measurements and discuss the uncertainties related to different measurements. The proportion of throughfall, evapotranspiration and runoff was 67%, 43% and 32% of the annual precipitation, respectively. The measured amounts of evapotranspiration and runoff were so small that the aim of closing the water balance of the studied system was not fully reached. The largest uncertainties are related to the evapotranspiration measurements and the determination of the actual surface area of the catchments used in the calculation of the runoff.

Introduction

In most terrestrial ecosystems the amount and seasonal variation in the availability of water is one of the key elements determining the typical biological processes, such as growth and survival, and species composition of the site. Depending on the location of the ecosystem, the key issue may be excess or shortage of water, either occasionally or continuously. In high northern or southern latitudes also the phase transition of water from liquid to ice can substantially affect the ecosystems and processes within them.

Boreal ecosystems are one of the largest biomes on Earth, making up ca. 20% of the total global forested area (FAO 2000). The fact that part of the precipitation falls in the form of snow and the variably long duration of the snow pack covering the ground and rapid snowmelt in springtime are important characteristics of the boreal forest water balance (Bonan *et al*. 1992). As a consequence of the climate change, changes in ecosystem water balances are expected to occur. In boreal ecosystems an increase in autumn and winter precipitation has been predicted, and correspondingly the amount of snowfall and amount of water stored in the snowpack are expected to decrease. Such changes in the timing and the amount of rainfall entering and leaving the system can change both the net primary productivity and decomposition, having thus also a feedback effect on the climate change itself. Water budget studies can provide baseline data on which carbon and chemical element cycling studies can be based (Luxmoore 1983).

In principle the composition of a water budget of a studied system is simple: measure and add together all components bringing water into the system, subtract all flow components out of the system and determine the changes in the system water storages. However, in practice this can be a very complicated task to carry out for a given area, and due to this multidiscipline, multiannual studies of ecosystem water budgets where all components of the water balance are measured simultaneously are rare (LaBaugh 1986).

The general equation of a water balance is as follows:

$$P + I = \mathrm{ET} + R + dW \quad (1)$$

where P = precipitation, ET = evapotranspiration, I = inflow, R = runoff, and dW = changes in water storages retained into the soil matrix and in the ground water. In any water balance study, it would be important that all these component fluxes were measured independently, and with equal precision. Often one or more of the measurements of the component fluxes are either very uncertain, inaccurate or not measured at all and estimated with models or with a subtraction method.

Taking into account the inaccuracies in the flow measurements, in large scale watershed studies, where all runoff occurs through the stream or river at the base of the watershed the runoff is known rather well through measurements of river discharge (Kirkby 1988, LeSack 1993, Hyvärinen and Korhonen 2003). At a smaller scale than watersheds, direct measurements of evapotranspiration may be available (by e.g. eddy covariance), but drainage and runoff are often unknown. Drainage can be estimated with pedo-transfer functions or as the difference between precipitation and evapotranspiration. Models are prone to uncertainty due to preferential flow caused by spatial variability in soil hydraulic parameters (Merz and Bárdossy 1998, Herbst and Diekkrüger 2002) and occurrence of macropores in forest soils (Bonell 1993, Mallants *et al*. 1998, Oliver and Smettem 2005). The reliability of the difference method relies on the accuracy of those measurements which are used to calculate the unknown flux. The number of sites where the evapotranspiration is measured continuously with eddy-covariance (EC) method, has increased rapidly (Baldocchi *et al*. 2001). The advantage of this method is its high temporal resolution, and the possibility to measure fluxes around the year, but there are only limited possibilities to estimate the accuracy of this measurement with comparisons to other means of evapotranspiration measurements.

A typical feature for the hydrological methods used in water balance studies is that one or even more components are calculated as residuals (Winter 1981). The value of monitoring the water balance components simultaneously at the same site is that the relations between the different measured water balance components can be studied in detail and the uncertainties of separate measurements can be evaluated by comparing the differences calculated between different measurements. For water balance studies, this gives the opportunity to check whether the measurements of water fluxes carried out with different methods are consistent with each other. If all fluxes are measured, it can even be checked whether the water balance can be closed. Climatic conditions occur quite randomly during different years and the variation between years in the amount and in the timing of the precipitation, evapotranspiration and outflow can be covered only if the measurements are continuous and the measurement period is long enough to cover different types of years.

At the SMEAR-II site, the components of the water balance have been measured continuously since 1997. In this paper, we report the annual

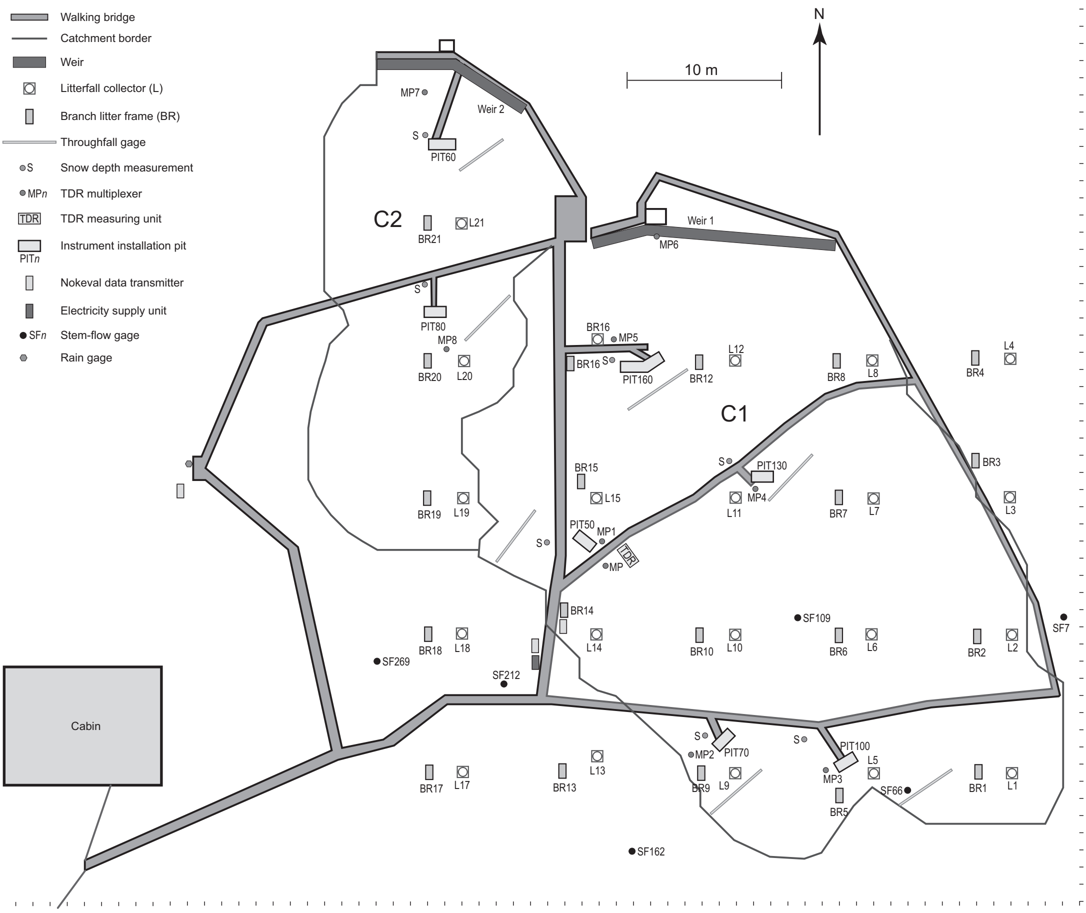

Fig. 1. Schematic picture of the measurement instrument installation in catchments. Borderlines of the catchments 1 and 2 are shown with a continuous line. The scale is shown with tickmarks of one meter interval around the figure.

values of all major components of the water balance of a young boreal Scots pine dominated forest. The objective of the study is to present the annual amount and variation in precipitation, throughfall, soil water content, runoff, transpiration and evapotranspiration. The sources and magnitude of uncertainties related to different measurements are discussed. We also test whether the water balance of the studied ecosystem can be closed using direct measurements.

Material and methods

Measurement station

The SMEAR-II station (**S**tation for **M**easuring **E**cosystem-**A**tmosphere **R**elations) was established in 1995 at the Hyytiälä Forestry Field Station to become a natural laboratory for studying material and energy balances and the processes underlying these balances between the forest and the atmosphere (Vesala *et al*. 1998, Hari and Kulmala 2005). The measurement station is located in southern Finland (61°51´N, 24°17´E, 180 m a.s.l.) in the boreal vegetation zone. The dimensions of the two micro-catchments (C1 and C2) at the SMEAR-II station and the instrumentation within them are shown in Fig. 1.

The annual long-term average temperature in the area is +2.9 °C; January is the coldest month (–8.9 °C) and July the warmest (+15.3 °C) (Fig. 2). The annual precipitation during the measurement period from 1959 to 2006 averages 697 mm (data from the Finnish Meteorological Institute (FMI) weather station at Hyytiälä).

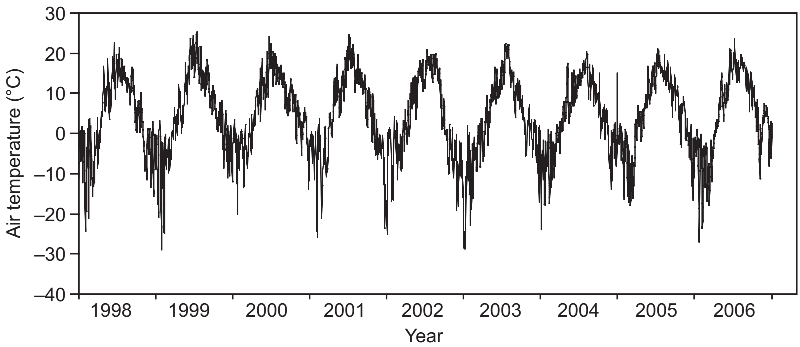

Fig. 2. Annual variation of the daily air temperature at the SMEAR-II station.

Precipitation measurements

The FMI weather station is located some 300 m west from the SMEAR-II station and is equipped with a manually-operated Tretjakov type rain gage installed at 2-m elevation above ground. The gage is equipped with a wind shield. The precipitation measurements at the FMI weather station were chosen to represent the above-canopy precipitation throughout this study. The estimated average annual correction factor of the precipitation for errors caused by evaporation, attaching, splashing, wind and installation location is assumed to be 1.18, varying from 1.07 in summer to 1.4 in winter (Kuusisto 1986: 35). Because of insufficient data for calculating the actual corrections for the daily values, the measured precipitation values were used and no correction coefficients were applied. However, in the water balance calculations it has to be taken into account that the actual amount of the annual precipitation has been higher than the applied values of the direct measurements. The amount of snowfall was estimated by summing up the precipitation during the days when the average air temperature was below zero.

The precipitation above the canopy was also measured at the measurement station with an automatic rain gage (ARG-100, Environmental Measurements Ltd.). The gage was installed on top of a 20 meter high measuring tower. The automatic measurements were running only during the frost free period of the year. These automated precipitation measurements at the SMEAR-II station were used only for comparisons with the measurements at the FMI weather station. The snowfall at the tower was measured with an open bucket method by melting snow and weighing the meltwater.

Throughfall measurements

We monitored the below-canopy throughfall at 0.7 m height above the soil surface with gages consisting of two stainless steel gutters with a total length of 4 m and a width of 0.1 m. An automatic tipping-bucket counter measuring the water running from the gutters was installed below the mid point of the two gutters installed in a gently sloping V-shape (Throughfall meter, Rainer, Pohja-Metallityöpaja, Juupajoki, Finland). The counter readings were recorded at 15-min intervals throughout the frost-free period of the year. The total open area of the gage was 0.385 m^2 and the amount of throughfall was calculated in mm (= l m^{-2}) by dividing the volume of the throughfall in liters by the open surface area of the wings of the gutters. Annual calibration constants for each gage were used for conversion. There was no shift in the calibration constant per year, and the variation between years was small (coefficient of variation 2%–8%). Similarly to the precipitation measurements, otherwise uncorrected values were used. The total number of the gages was seven, out of which 5 gages located in C1 and two gages in C2. The tipping bucket counter values were also controlled by collecting the water running through the counters into 20 liter containers which were weighed with varying intervals to give the cumulative amount of the throughfall during the period between weighing.

The throughfall measured from each gage was averaged for each day, separately for both catchments. The values of the seven throughfall measurements were cross-checked to screen for erroneous readings. For the years 1999, 2000, 2002 and 2005, one counter (a different one each year) gave biased results and the data were therefore rejected. The interception of the ground vegetation was not measured.

Stemflow was measured with four stemflow collectors made of a cleaved silicon rubber tube (25 mm in diameter) which was installed water-tightly with a silicon seal around (> 360°) the trees at 100 cm height. The stemflow was lead from the collectors to 12-liter containers, which were emptied and weighed on a weekly to monthly basis (depending on the amount of rainfall) during the snow free period. The amount of stem flow (mm) was calculated from the average amount of water (l) collected annually from the stem flow collectors multiplied by the amount of trees at the site (Table 1).

Snow measurements

The snowfall below the canopy during winter was measured from 7 snow collectors (50 cm in diameter) between 1998 and 2000 and from 2 collectors between 2001 and 2006. The melted water was weighed on a monthly basis throughout the winter. The snow depth was measured on a weekly basis from 7 snow depth measurement poles located close to the throughfall gages. During the winter of 2004, the water content of the snowpack (cm cm^{-1}) was measured with regular intervals. For estimation of the water balance during the snowmelt periods the change in the water storage of the snowpack needs to be estimated. A good positive relationship was found between the snow water content and a moving average of the air temperature over the last 7 days (Fig. 3). The change in the storage of water in the snow was estimated using the snow depth measurements and this relationship.

The dynamics of the snowpack was also estimated with measurements of snowfall from the FMI station by using a simple accumulation and melt model: The precipitation was assumed to be snow when the daily average temperature was below zero, and snow accumulation occured during these periods. The snow melted at a constant rate (mm °C^{-1} day^{-1}) when the temperature was above zero. The best fit with the disappearance of the snowpack at the SMEAR-II station was found to be 1.5 mm °C^{-1} day^{-1}. Several threshold values for the snowmelt were tested, and the 0 °C threshold value was selected.

Using the melting from the simulated snowpack ensures that the total amount of water eventually released from the snowpack is equal to the measured snowfall from the FMI station. Using the snow depth directly would be much more inaccurate, because the snow water-content is highly variable, and fluctuations in snow depth do not necessarily represent the water released from the snowpack.

Soil water storage and runoff measurements

The catchments were established on a top of a hill in order to enable a selection of depressions

Table 1. The stand properties of catchments C1 and C2 before and after the thinning of C2.

	2001		2003	
	C1	C2	C1	C2
Avgerage diameter at 1.3 m height (cm)	12.5	10.7	12.7	11.3
Avgerage diameter at 6 m height (cm)			10.5	8.7
Height (m)	13.1	11.1	13.9	11.8
Total number of trees	167	52	153	33
Number of trees per ha	1879	1728	1721	1096
Average volume of a tree (l)	81.5	51.7	88.3	61.4
Volume of stems (m^3 ha^{-1})	153.1	89.3	151.9	67.3
Basal area of stems (m^2 ha^{-1})	23	15.3	21.7	10.9
Needle biomass (kg ha^{-1})	6370	4570	5970	3180

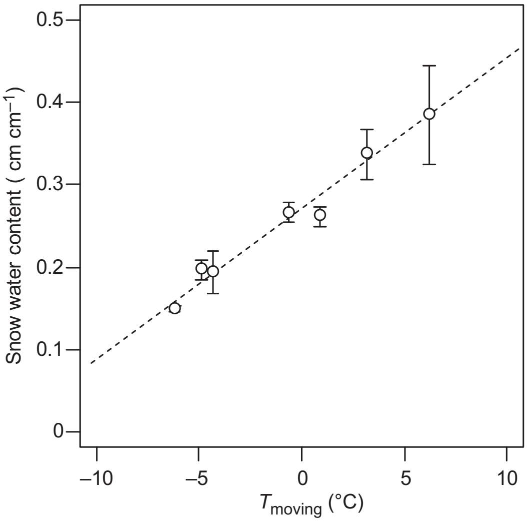

Fig. 3. Measured snow water-content during the winter of 2004 as a function of the moving average of air temperature over the last 7 days. Measurements were replicated at seven locations for each date, the bars denote standard errors.

forming micro-catchments with known borders and to avoid all unknown lateral water flow from surrounding areas. During the establishment of the measurement station the land area was divided into 1 × 1-m grids, and the elevation at each corner of the grid was measured to produce a contour map of the soil surface. The soil depth at these same points was measured with a soil radar. The radar signal was calibrated with actual soil depths measured from seven excavated pits and at the weirs. The contour map of the bedrock was based on the information of elevation and soil depth. The radar signal was also used to reveal possible cracks in the bedrock ensuring that no uncontrolled leakage of water out of the system occurs. No major vertical cracks were found, suggesting that no leakage through the bedrock exists.

The surface shape of the bedrock forms two adjacent basins (C1 and C2) that can be regarded as separate hydrological units. When measured by the grid method, the surface area of the larger basin (C1) was found to be 889 m^2 and that of the smaller (C2) 301 m^2 (Fig. 1) These catchments receive water only in precipitation. The soil covering the bedrock at an average depth of 50–70 cm is Haplic podzol formed on glacial till (FAO-UNESCO 1990). The average humus layer depth is 5 cm. Some general properties of the soil are presented in Table 2.

The total soil volume of catchments C1 and C2 was determined using the soil depth measured at each corner point of the grid. The estimated total soil volumes of C1 and C2 are 541 m^3 and 151 m^3, respectively. The estimated soil volume is also divided into different soil horizons (Table 2). The bulk of the soil volume is in the B-horizon, which has here been defined as a layer at 9–22 cm depth. The shape of the bedrock is such that the proportion of the deep-

Table 2. The soil properties in catchments 1 and 2. Soil layers were defined by their top and bottom distances (cm) of from the soil surface. The layer 0–5 cm is the humus layer.

	0–5	5–9	9–22	22–34	34–55	55–72	> 72	Total
Catchment 1								
Stone content (% of volume)	0	26	28	31	31	43	43	
Soil volume (m^3)	44.5	34.8	112.6	100.7	147	72.4	28.6	540.7
Soil volume without stones (m^3)	44.5	25.8	81.1	69.5	101.4	41.3	16.3	379.8
Water in soil (m^3)								
all pores filled	28.0	17.0	38.1	32.7	39.6	16.1	6.4	177.8
at field capacity (pF2)	14.7	8.8	24.3	20.8	26.4	10.7	4.2	110.0
at wilting point	3.1	2.1	4.9	4.2	8.1	3.3	1.3	26.9
Catchment 2								
Stone content (% of volume)	0	26	28	31	31	43	43	
Soil volume (m^3)	13.6	13.6	33.5	28.2	44.7	15.1	2.7	151.4
Soil volume without stones (m^3)	13.6	10.0	24.2	19.5	30.9	8.6	1.6	108.3
Water in soil (m^3)								
all pores filled	8.5	6.6	11.6	9.2	12.1	3.4	0.6	51.7
at field capacity (pF2)	4.5	3.4	7.2	5.9	8.0	2.2	0.4	31.7
at wilting point	1.0	0.8	1.5	1.2	2.5	0.7	0.1	7.7

est soil horizons of the total soil volume is small. The amount of stones in each soil layer of each pit was measured by weighing, and the proportion of the stone volume of the pit volume was estimated (stone density 2.65 g cm^{-3} was used).

The downslope side of the catchments was excavated down to the bedrock. The bedrock was washed free of soil and the bases of two separate weirs were cast to the bedrock with water tight concrete. Further, a water-resistant plywood reaching the top of the soil was attached to the concrete base to form a water tight wall preventing all leakages from the catchments. A runoff tube with a diameter of 5 cm was installed in the lowest point of the weirs of both catchments, and the tubes were connected to flow meters (Schlumberger Aquatic, Schlumberger Water Services). For the measurement of the surface runoff at the weir, L-shaped stainless steel plates were pushed into the eluvial layer soil, and a tube for runoff was installed in order to collect and measure the surface flow with similar flow meters as have been described above. The runoff was measured in liters which were transformed to mm values by dividing the amount of outflow in liters with the estimated surface area of the catchments in square meters. The cumulative flow readings were also recorded manually at 7-day intervals to check the validity of the automated measurements. The flow meters were calibrated annually in the laboratory by introducing a known amount of water through the meters at different flow rates. The installation of the weirs and other equipment was started in May 1995 and finalized in August 1995.

The soil water retention curves of the different soil horizons were determined by the pressure plate method from soil cylinders sampled from the walls of the seven pits. Based on the soil water retention measurements the hydraulic parameters such as total soil air space, soil water content at field capacity and at wilting point were calculated (Table 2).

The soil volumetric water content was measured continuously with time-domain reflectometry (TDR) by using Tektronix 1502 C cable radar (Tektronix Inc.) between 1998 and 2004 and from 2005 onwards with TDR100 (Campbell Scientific Inc.). The probes were connected with a coaxial cable (type RG 58) to multiplexers (SDMX50, Campbell Scientific Ltd.) and further to a data logger (Campbell 21X, Campbell Scientific Ltd.). The TDR-probes consist of two rods of stainless steel (175 mm long, 5 mm in diameter). A total of 64 TDR probes were installed and they were distributed in different soil horizons of the 7 soil pits and in the soil adjacent to the weir. In the calculations the values of 39 sensors distributed in the area of the catchments were used and the values from the sensors located at the edge of the weirs were omitted in order to avoid giving too much weight for the possibly altered moisture conditions at the weir. We used the Ledieu *et al*. (1986) calibration for calculating the volumetric water content:

$$\theta_{\mathrm{v}} = a\sqrt{K_{\mathrm{a}}} - b \tag{2}$$

where θ_{v} is the volumetric water content (m^3 m^{-3}), K_{a} is the apparent dielectric constant, and a and b are parameters from Ledieu *et al*. (1986) for mineral soil. For the humus layer, we used a and b parameter values presented in Pumpanen and Ilvesniemi (2005).

Due to the elevated location as compared with surrounding areas and the shallow soil there is no actual groundwater in the soil of the catchments. However, in order to determine conditions when the soil water content was over the field capacity, the soil moisture contents measured by TDR were compared with soil water retention (pF) curves. The maximum soil volumetric water contents measured with TDR gave higher soil volumetric water contents than the pF curves at the 0-tension level. Despite this disagreement, no correction for soil water contents measured by TDR were made for the calculations of the total volume of water stored in the soil of the catchment. The TDR measurements were done at hourly intervals throughout the year, but a 24-hour average was calculated from the raw data to reduce the noise in the measurement signal. All values presented in this study are daily averages.

The soil water storage was estimated from the measurements of the soil water content for each soil layer in both catchments as follows:

$$W_i = \frac{1000\theta_i V_i \left(1 - \rho_{\mathrm{R}}\right)}{A} \tag{3}$$

where W_i is the water storage (mm) for layer i, θ_i is the volumetric water content (m^3 m^{-3}) for layer

i, V_i is the total soil volume of the layer i (m^3), A is the soil surface area of the two catchments, and ρ_R is the stone percentage of the layers. The total water storage in all soil layers together was calculated by summing the values of all soil layers.

Evapotranspiration

Estimates of the evapotranspiration were obtained with the eddy covariance (EC) method. A detailed description of the instrumentation and installation is given by Vesala *et al*. (1998). In short, the EC measurements were conducted at a flux tower at 50 m distance from the catchments. The wind speed was measured by sonic anemometers (Solent Research 1012R2, Gill Instruments Ltd.). The H_2O concentrations were measured by an infrared absorption gas analyser LI-6262 (Licor Inc.). The measuring height of the fluxes was 23 m approximately 10 m above the forest canopy except from April 1998 until June 2000 when the fluxes were measured at 46 m (Vesala *et al*. 2005). The sample line was 7-m long and the outside/inside diameter of the tube was 6/4 mm. The material of the tube was PTFE from the start of the measurements until May 2002. On 9 May 2002 the sample line was renewed and the old tube was replaced with electro polished seamless stainless steel tube of the same diameter. The effect of the tube on the results of EC measurements is described in detail in Mammarella *et al*. (2009). The measured annual sum of ET corrected with a constant response time gives about 10% lower values than the one corrected with the effective response time.

The collected data were quality controlled and corrected for frequency losses and sensor separation according to standard procedures (Rannik 1998, Aubinet *et al*. 2000). The EC fluxes were calculated as 30-min block-averaged co-variances between horizontal and vertical wind velocity.

The footprint of the EC measurements is larger than the catchments, and although the area is mainly covered by pine stands of similar age, the vegetation is not totally homogeneous within the footprint area. There is also a steep descending slope westwards from the tower. The footprint also depends on the wind speed. Rannik *et al*. (2006) estimated that the uncertainty in the net ecosystem exchange of CO_2 at this site is about 80 g C m^{-2} $year^{-1}$, or 30%–50% of the total. However, the uncertainty in evapotranspiration may not be directly linked to that of CO_2. The surface energy balance closure is typically imbalanced in a way that the sum of the sensible and latent heat fluxes is less than the net radiation and the average bias is in the order of 20%. It is not known how the imbalance of 20% is distributed between sensible heat flux and evapotranspiration, but conservatively we can conclude that the underestimate of evapotranspiration by 20% is not atypical, resulting possibly from the variation of the flux field over complex terrain, advective transport and convective cells (Foken 2008).

The evapotranspiration at the soil surface was measured continuously with 2–3 automated transparent chambers (d and h = 20 cm). The chamber has a lid that closes automatically once an hour for the measurement period of 70 s. A continuous air flow was directed through the chamber to the automatic gas exchange system of the SMEAR-II station. The technical details of the chambers are described in Pumpanen *et al*. (2001). The soil H_2O efflux was estimated by integrating the results of the direct chamber based measurements over the period considered. Until the year 2000 the vegetation had been removed from the chambers, but later the ground vegetation in the chamber was left untouched. However, the amount of plants in the chamber was lower than in the area in general. Since 2004 the locations of chambers have varied.

The transpiration of the pine trees was determined by integrating chamber-based H_2O fluxes of three to four shoot chambers. The methods of the H_2O flux measurements as well as the spatial and temporal integration (half hour steps) of the fluxes have been documented in detail in Kolari *et al*. (2009).

Forests on the EC footprint and catchments

The dimensions of the standing trees in the 200-m radius area around the EC mast were measured in 2001 and the succeeding years.

Systematic plot sampling along radial directions starting from the mast was used. The circular sample plots (100 m^2) were located on 16 radials whose central point was the mast. The plot interval was 20 m, and the centre of the first plot was located 5 meters from the mast in the N and S radials, 10 m in the E and W radials, 25 m in the NE, SE, SW and NW radials and 45 m in the 8 other radials. These eight radials were measured only in 2001. The variables of interest were the total tree biomass and the biomass of the stems, branches, needles and roots.

The dominant tree population on the site is a 45-year-old Scots pine (*Pinus sylvestris*) stand, which was established by sowing after prescribed burning in 1962. The number of pine trees in 2006 in the 200 m radius footprint area around the EC tower was 960 ha^{-1} and the average height and diameter at 1.3 m height of pines were 15.5 m and 14.1 cm, respectively. In addition there were approximately 1100 per ha small spruces (average height 6.8 m and average diameter 3.9 cm) and 3600 per ha broadleaved trees (average height 5.5 m and diameter 1.8 cm). In the winter of 2001–2002, a section of the 200 m radius footprint area (4.3 ha) was thinned and the total amount of stem volume removed in the thinning was 141 m^3.

The properties of the forests growing in C1 and C2 are shown in Table 1. The stand in C2 was thinned in the beginning of 2003. The thinning decreased the number of trees per hectare from 1728 to 1096 (–38%). The calculated amounts of the needle biomass in C1 and C2 after thinning were 5970 and 3180 kg ha^{-1}, respectively, and the reduction in needle biomass in C2 was 1400 kg ha^{-1} (30.5%) (Table 1).

The dominant species in the understorey vegetation are *Vaccinium myrtillus* and *Vaccinium vitis-idaea*. The ground vegetation consists mainly of mosses *Dicranum polysetum*, *Hylocomium splendens* and *Pleurozium schreberi*.

Estimating the water balance from measured fluxes

The hydrological year was defined as the calendar year (1 January to 31 December), because the autumn rain episodes filled the soil water storage to the field capacity and the outflow ceased due to below-zero temperatures before the end of the year. The changes in soil water storage determined from TDR (ΔW_{TDR}) was calculated on an annual basis as follows:

$$\Delta W_{\mathrm{TDR}} = W_{\mathrm{TDR}}(t_{i+1}) - W_{\mathrm{TDR}}(t_i) \qquad (4)$$

where $W_{\mathrm{TDR}}(t_i)$ is the water storage in the beginning of the year (mm) and $W_{\mathrm{TDR}}(t_{i+1})$ is the water storage in the end of the year (mm). Because the TDR primarily measures the dielectricity of the soil, and because the dielectric constant of ice is very different from that of liquid water, we excluded the TDR readings of frozen soil and assumed that no changes in soil water content occurred when the soil was frozen. The last unfrozen reading of the preceding year was used as a starting point for the soil volumetric water content change calculation in the succeeding year.

The ΔW_{TDR} was compared to the ΔW determined for selected periods from the different components of the stand water balance as follows:

$$\Delta W = P(t) - E(t) - R(t) \qquad (5)$$

where ΔW is the change in soil water storage (mm), P is the precipitation above the canopy (mm), E is the evapotranspiration (mm), R is the runoff (mm), and t = time. Because the measurement of E includes the evaporation of canopy intercepted water, P is the precipitation above the canopy, not precipitation that reaches the soil surface (= throughfall).

Results

Precipitation

The annual precipitation during the eight years studied varied from 535 mm in 2002 to 825 mm in 1998. The average precipitation during the period between 1998 and 2006 was 692 mm (Table 3). The amount of annual snowfall (rainfall during days, when air temperature was below zero) varied between 113–250 mm and the proportion of snowfall in annual precipitation was between 18%–46%. The timing of the

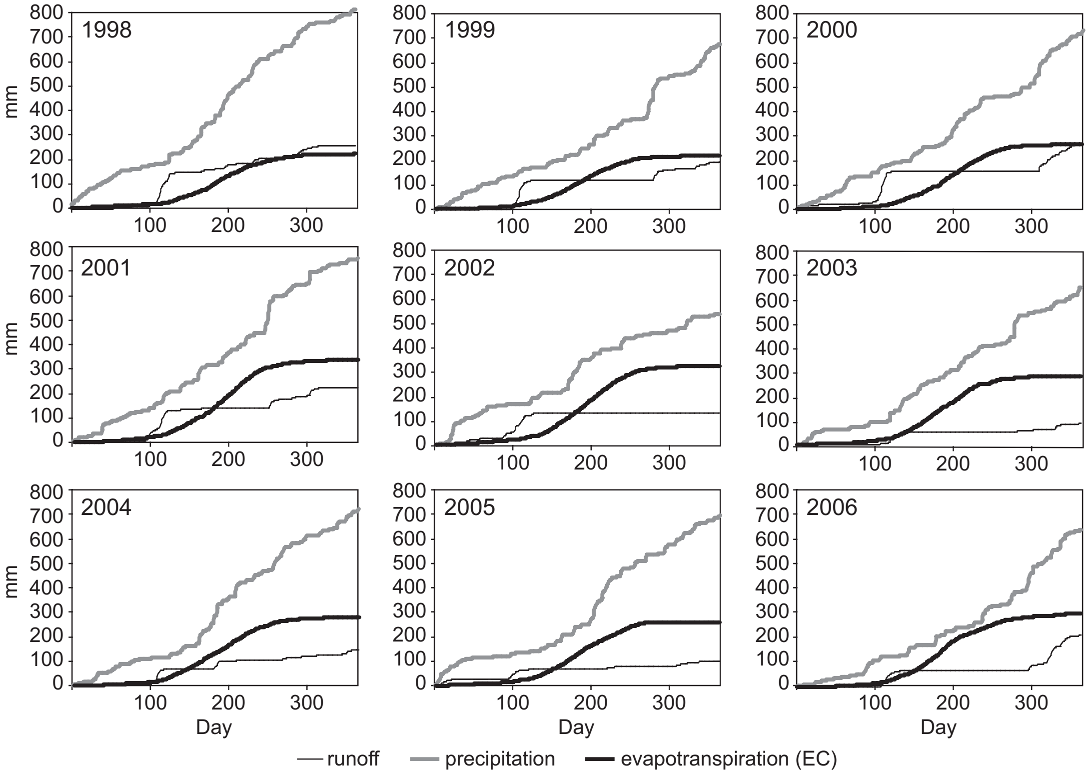

Fig. 4. The cumulative precipitation, evapotranspiration (EC) and runoff in years 1998–2006.

rain episodes within the year varied, but typically the increase in the cumulative precipitation was smallest in the winter months, and the periods with the highest precipitation rates in autumn (Fig. 4).

The automatic rain gage installed above the canopy showed the same daily peaks as the manual measurements, but due to different reasons (e.g. lightning and maintenance) causing breaks in electricity supply, some rainfall events

Table 3. Annual values (in mm) of the components of the water balance of the catchments at SMEAR II station and the footprint of the eddy covariance measurements. P = precipitation, S = precipitation during days when air temperature has been below zero, an estimate of the amount of snowfall, T = throughfall, I = canopy intreception, R = runoff, Tr = transpiration, ET = Evapotranspiration, measured by eddy covariance (EC) or difference method ($P - R_{C1+C2}$), C1 and C2 = catchments 1 and 2, R_{C1+C2} = sum of the runoff in liters from C1 and C2 divided by the total area (m^2) of C1 + C2.

Year	P	S	T	I	R_{C1}	R_{C2}	R_{C1+C2}	Tr	ET (EC)	ET ($P - R_{C1+C2}$)
1998	825	229	549	276	234	368	268	122	235	557
1999	676	251	537	139	177	245	195	157	233	481
2000	730	164	490	241	245	329	267	130	270	463
2001	752	178	511	242	205	288	227	147	356	525
2002	535	244	340	195	122	154	130	175	340	405
2003	645	149	385	260	91	199	119	141	320	526
2004	718	186	445	274	139	264	171	146	294	547
2005	698	196	460	238	122	222	147	160	–	551
2006	644	113	451	193	213	330	242	163	313	402
Average	692	190	463	229	172	267	196	149	295	496

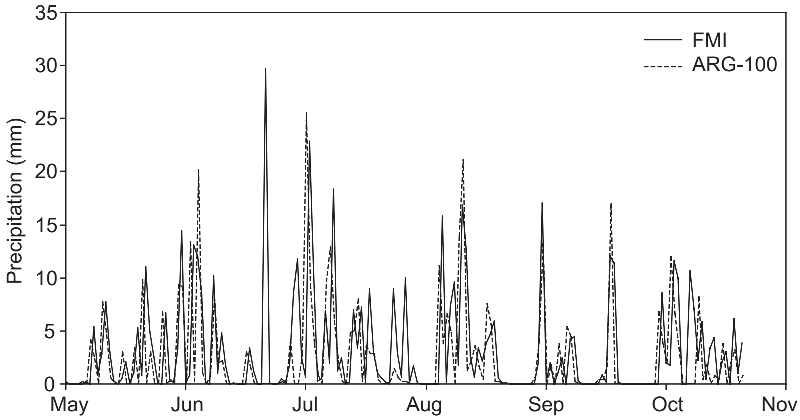

Fig. 5. The timing and amount of rain episodes measured by automatic ARG-100 gage and manual measurements between May and October in a rainy year 1998.

were not detected (Fig. 5). During the frost-free period, when the automatic measurements were running, the automatic gage gave somewhat lower (average of all years 10%, range 5%–14%) precipitation values than the FMI station. Part of this difference was caused by the missing events, but also on the days when measurement unit was operating correctly (which was most of the days), the precipitation measured with the automatic gage gave lower values than the manual measurements.

The timing of the rain events between these two measurements was not exactly the same since the automatic measurements represent the values of the actual 24 h of a calendar day and the FMI measurements were conducted daily at 8 am and only from Monday to Friday. This means, that if a rain event occurred e.g. on Saturday, the FMI record shows the value on the following Monday.

Throughfall

The annual amount of throughfall varied between 340 and 549 mm (Table 3). The amount and timing of the throughfall measurements during the period when automatic measurements were carried out fit well with the above canopy precipitation measurements (Fig. 6). The proportion of canopy interception calculated on annual basis was on an average 33% of the precipitation. Between June and September this proportion was 37%. The proportion of the canopy interception is higher when the rainfall intensity is low (Fig. 7). When catchments C1 and C2 are compared, it can be seen that the amount of the throughfall was higher in C2 and the difference between the catchments increased after the thinning (Table 4). The standard error of the mean in the throughfall measurements varied between 5–15 mm in different years. The late autumn and wintertime throughfall in the years 1998, 1999 and 2000, when 7 collectors were available were 208, 216 and 219 mm, respectively. The amounts of precipitation during the same years were 237, 261 and 266 mm, respectively, giving an average wintertime throughfall proportion of 84%.

The annual amount of stem flow varied between 3 and 10 mm with an average of 8 mm. The stem flow forms only a minor proportion of the water flows of the catchments.

Water storage in snow

The amount of precipitation during the days when the air temperature was below zero was on average 190 mm and varied between 113 and 251 mm (Table 3). The maximum water storage in the snowpack, measured from the depth and the average water content of snow, ranged from 100 mm to 150 mm (Fig. 8). Thus the maximum snow storage is smaller than the snow deposition in open areas (Table 3). The variation between measured snow depths was high in autumn and spring when the proportion of the standard error of the mean of the average varied between 15%–70%, whereas during the deep snow period it was only 5%.

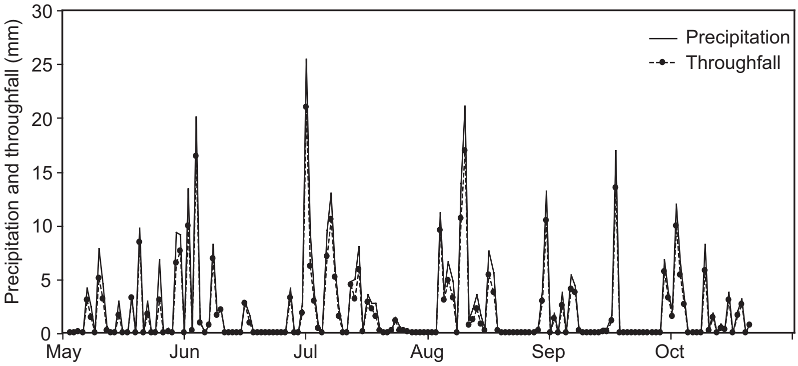

Fig. 6. The timing of rain episodes and the amount of throughfall and precipitation in 1998.

In most of the years, the snow water storage started to accumulate in December and peaked in late March. The only exceptions to this pattern were the years 2000 and 2003, when the first snow came in late October. The inter-annual variation in the snow water storage was rather small as compared with the soil water storage capacity. In spring, the thawing of the snow was very rapid and took place in late April and early May.

A comparison between the simulated snowpack and the estimated storage from snow depth shows a good agreement, not only in the timing of the snowmelt, but also in total storage during the winter (Fig. 8).

Soil water storage

There was a clear seasonal pattern in the soil water content and consequently, soil water storage. The soil water storage was at its highest in April during the snowmelt and in November and December after the autumn rains, when the soil in the deep layers was saturated (> 50% volumetric water content) and near field capacity in the surface layers (20%–30%). The lowest soil water content values were measured in late July and August, 20% in the deepest soil layers and 15% in the humus (Fig. 9). The standard error of the mean between the TDR probes installed in the same soil horizon varied between 2%–4%. The maximum soil water storage in late autumn when the outflow had ceased was around 270 mm in both catchments. The lowest summertime soil water storage during the measured years was lower than 100 mm. In dry periods during summer the decrease in soil water storage was continuous and the rate of change rather constant. The soil in C2 did not dry as much as the soil in C1 (Fig. 10). This different rate of soil drying between the catchments correlates with the differences in the amount of needle biomass

Table 4. The annual throughfall (mm) measured during frost-free periods in catchments 1 (n = 5) and 2 (n = 2). 'C2 – C1' is the difference of the cumulative throughfall between the catchments. 'C1 SE' is the standard error of the mean of 5 counters in catchment 1, and 'C2 difference' is the difference between the two counters in catchment 2.

Year	Average in C1	Average in C2	C2 – C1	C1 SE	C2 difference
1998	362	393	31	19	51
1999	306	381	75	37	207
2000	316	325	9	26	16
2001	302	309	7	20	49
2002	144	156	12	10	35
2003	220	282	62	8	80
2004	310	346	36	10	20
2005	272	300	28	6	50
2006	244	268	24	5	9
1998–2002	286	313	27	22	72
2003–2006	262	299	38	7	40

between the catchments (Table 1).

Small rainfall episodes did not change the measured soil water content. After some successive rainfall episodes, the soil water content increased and the amount of change in the soil water storage was in the same range as the measured throughfall. In autumn (October–November) the soil water storage started to fill up reaching its maximum capacity before the permanent snowpack was formed. Due to this, no major changes in the annual water storage were found during the measurement period, except in 2002, when the soil water storage did not reach its field capacity at the end of the year.

The amount of precipitation needed to start an outflow from the weir and the associated time-lag between the beginning of a rain episode and an outflow is shown in Fig. 11. The year 2000 was selected as an example year, because autumn was dry, and the soil water storage continued to decrease until the end of September. After the dry period the soil water storage in C1 was 150 mm. During the period between 1 October and 7 November, when the outflow from the weir started, the amount of precipitation was 130 mm and the measured throughfall 80 mm. When the outflow began, the amount of water stored in the soil was 270 mm (Fig. 11). Because the amount of water needed to increase the soil water content from 150 mm to 270 mm is almost equal to the amount of precipitation, this suggests that the throughfall measurement must be an underestimation. The time lag from the start of the rainy period to the beginning of the outflow at the weir was 36 days. There was a positive correlation between the daily precipita-

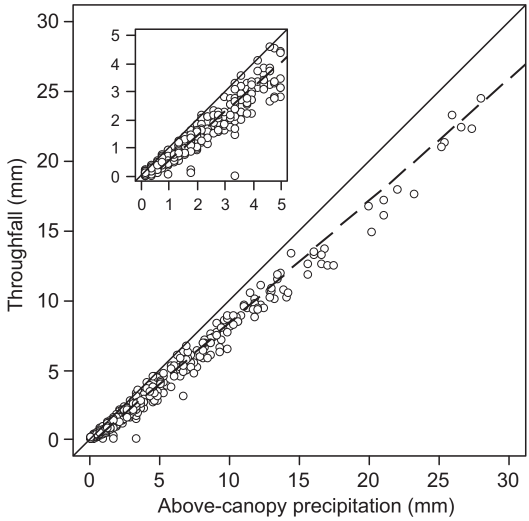

Fig. 7. Comparison between daily measured precipitation above canopy (ARG-100) and throughfall (average of seven gages) during the snow-free period. The insert shows the same data separately for the small rain events (rainfall during 73% of all rainy days was less than 5 mm). Solid line = 1:1 ratio; dashed line = linear regression: throughfall = –0.327 + 0.877 × precipitation (R^2 = 0.97, p < 0.0001). On average, all rain was intercepted when the daily rainfall was less than 0.4 mm.

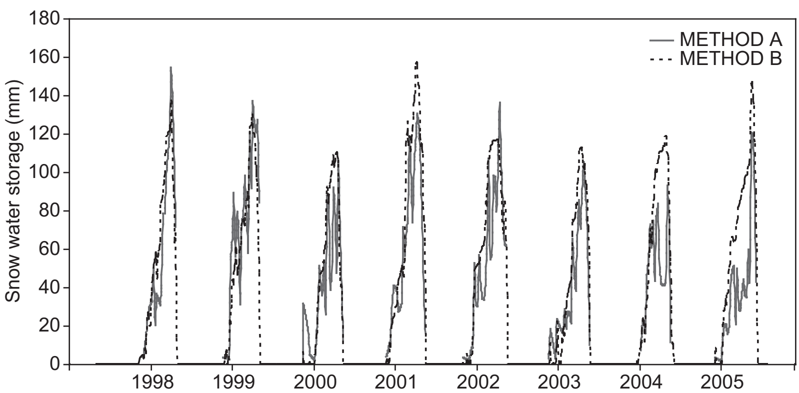

Fig. 8. The annual accumulation and melting of the water storage in the snow-pack. Snow water storage was estimated by two methods. Method A: (bi-)weekly measurements of snow depth at seven locations at the SMEAR-II site were converted to snow water storage using a temperature-dependent estimate of snow water density. Method B: snow water storage was estimated from measured precipitation at the FMI weather station, using a simple accumulation–melt model. The numbers on x-axis denote the end of the year.

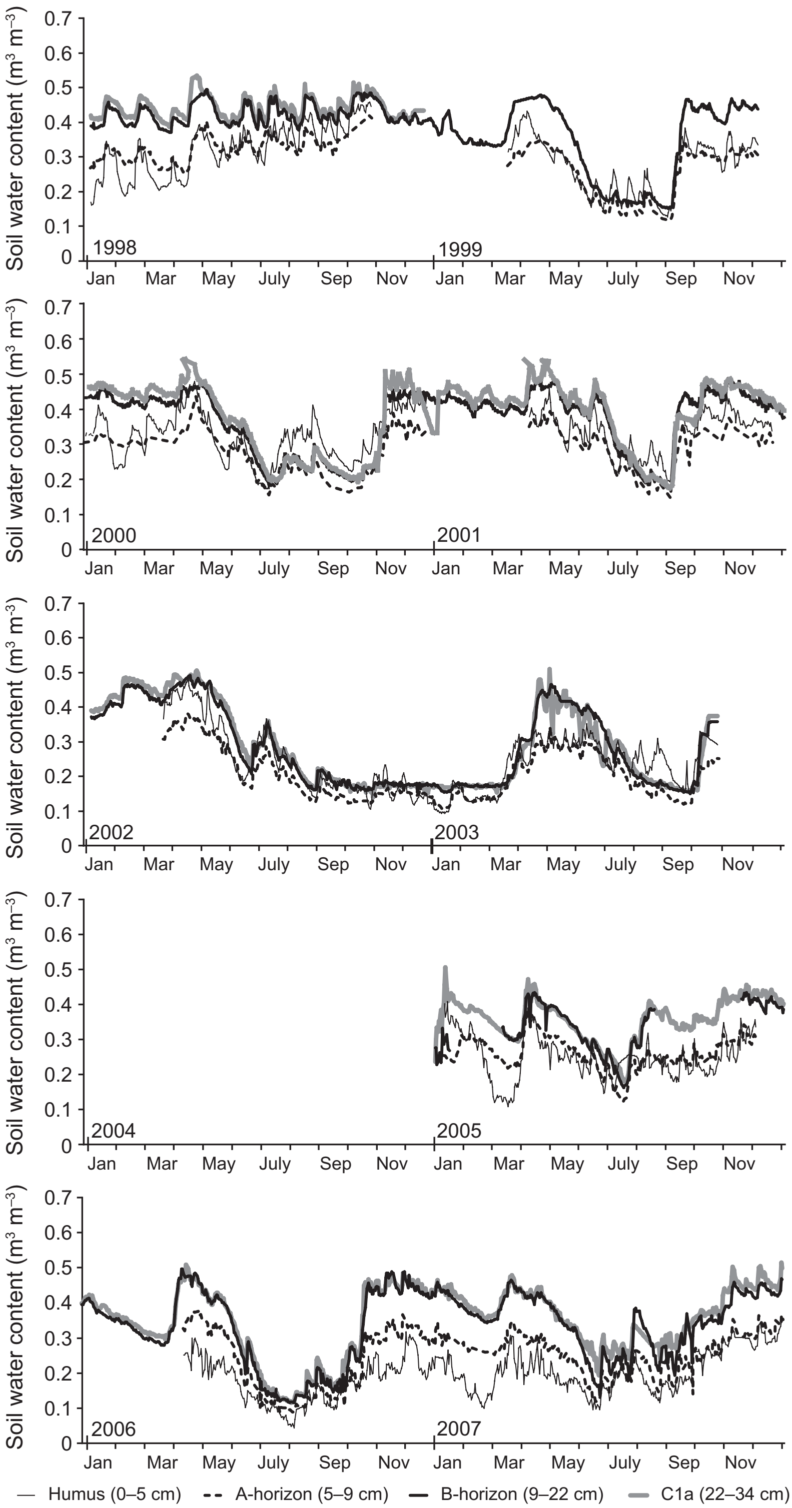

Fig. 9. Volumetric water content of the catchment 1 at four soil depths. The water contents are measured with TDR method. The results are averages of 5 soil pits (in total 31 sensors). In 2004, the signal from the TDR probes was too noisy and all the TDR data of that year were rejected.

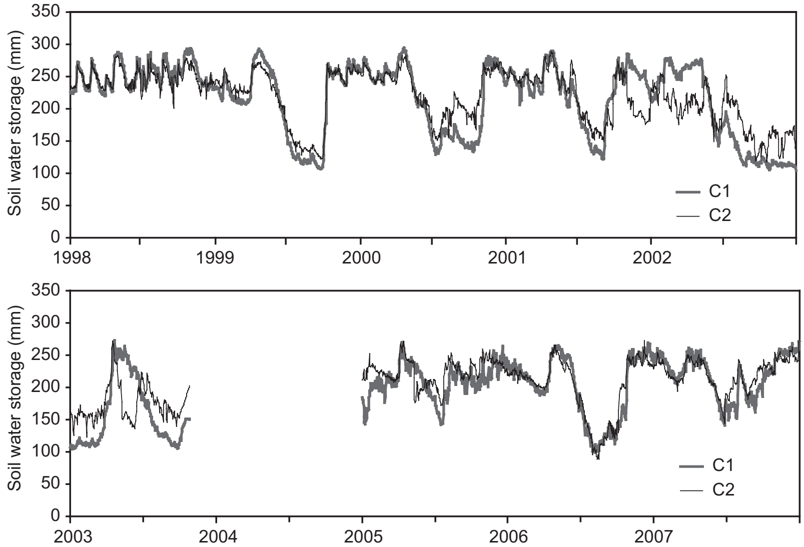

Fig. 10. Variation in the soil water storage in 1998–2006 in catchments C1 and C2.

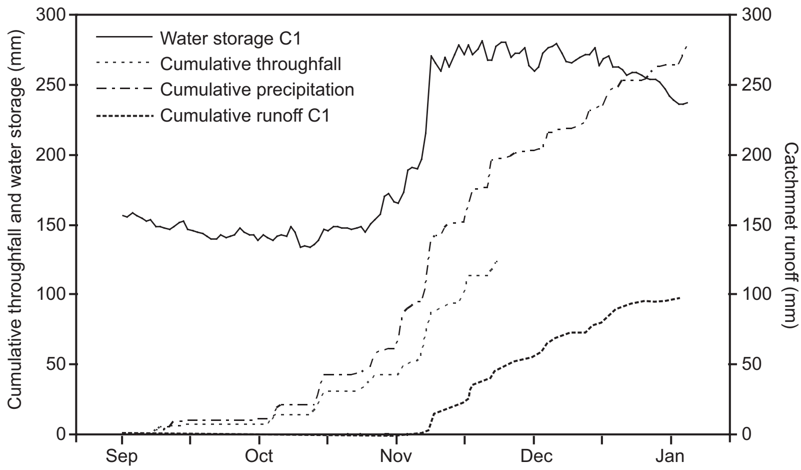

Fig. 11. The timing of precipitation, throughfall, change in soil water storage and runoff in catchment 1 in autumn 2000.

tion and outflow measured from the weir during the time when the soil water content was already at field capacity, but the coefficient of determination was rather low ($R^2 = 0.52$).

Runoff from the catchments

In every year, there was runoff from both catchments. There was a large inter-annual variation in the amount of runoff, but in most years the timing and variation in runoff in both catchments was similar (Fig. 12). The annual average runoffs measured between 1998 and 2006 were 172 mm and 267 mm in C1 and C2, respectively (Table 3). The average runoff from both catchments during the study period was 193 mm.

Most of the runoff from the catchments took place during the short snowmelt period in late April and early May (insert in Fig. 12). Another, but smaller runoff peak occurred in October and November. Runoff during the summer was not common and took place only in the years 1998, 2001 and 2004 when the deep soil horizons were saturated with water after a period of successive rain episodes. The difference in the runoff between the catchments increased in very rainy years and consequently decreased when the year was dry. In the exceptionally dry year 2002, the difference in the runoff between the catchments

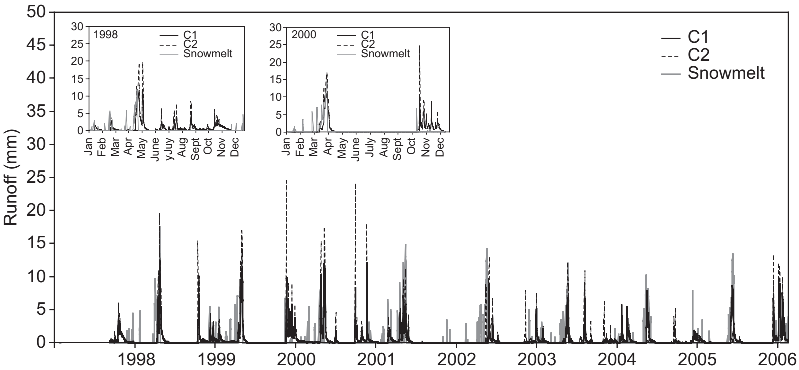

Fig. 12. Measured runoff of the catchments C1 and C2. The insert shows a magnification of the spring snow melt in 1998 and 2000. Note that a daily runoff of 10 mm corresponds to 8890 liters per day in catchment 1 and 3010 liters per day in catchment 2. The values on the *x*-axis denote the end of the year.

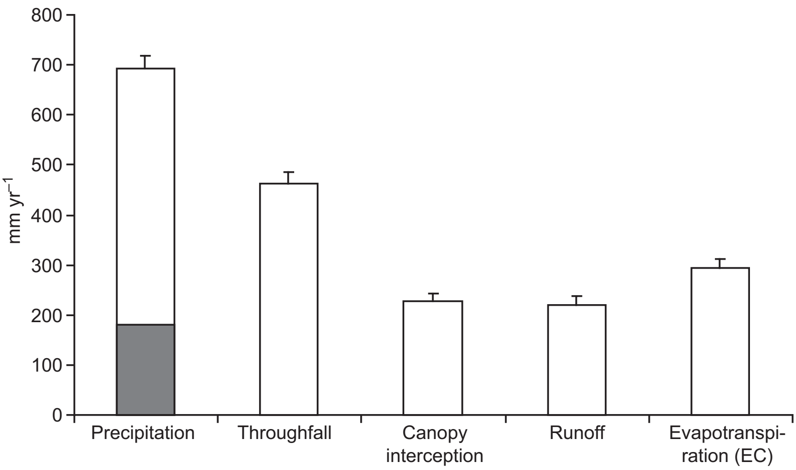

Fig. 13. The average values of the components of the water balance over the whole study period. The proportion of precipitation in snow has been marked in the precipitation bar.

was 32 mm, whereas in the year 1998 it was 134 mm.

We also measured the surface runoff under the humus layer, but surface flow never took place indicating that the water first moved vertically to deeper soil layers during spring or during heavy autumn rains and the lateral flow started near the bedrock when the field capacity of these layers was exceeded.

Evapotranspiration

We estimated the evapotranspiration with two independent methods, by eddy covariance measurements and by the difference method between the precipitation and the runoff. The evapotranspiration determined with the eddy covariance method was on average 295 mm and varied between 233 and 356 mm (Table 3).

The evapotranspiration determined as a difference between precipitation and average runoff of both catchments was 472 mm and varied between 402 and 557 mm. So, the difference between water balance and EC evapotranspiration estimates was 177 mm. If we calculate the difference between runoff and precipitation separately for both catchments, the estimate of annual evapotranspiration varies between 413 and 591 mm in C1 and between 314 and 477 mm in C2. The annual average of the transpiration of the pine trees was 149 mm and it varied between 122 and 175 mm (Fig. 13).

The difference in precipitation and EC determined evapotranspiration and runoff measured at the weirs gives an offset of 217 mm in the C1 and 123 mm in the C2. When this difference was compared with the changes in soil water storage measured with TDR the two methods showed systematic behavior, but there was an offset of 100 mm in the constant of linear equations fitted to the results of both catchments (Fig. 14). During the summers (1 June–31 August) of 1999–2005, the evapotranspiration calculated as a sum of the change in soil water storage and precipitation was 241 mm as compared with 149 mm of the EC evapotranspiration.

Discussion

Precipitation

The measurement period of this study covers one of the driest and wettest growing seasons of the measured history in the area. This allows us to analyze the behavior of the ecosystem water balance in climatic conditions which probably cover the variation that can be expected to be found also in the near future.

The manual precipitation measurements were shown to be less prone to technical problems and were therefore taken as a reference measurement. It seems that the reliability of the automatic gage is not as good as manual measurements, since every season some rain episodes were lost, suggesting that when automatic gages are used, parallel manual measurements should be carried out to backup the possible gaps in the data. There was also a small, offset type of difference, as the values otherwise measured correctly, gave slightly lower values than the manual measurements. The automatic gage was installed above the canopy and did not have a wind shield. The wind speed above canopy and at the 2-m elevation can differ and because the results were not corrected for wind this can be one possible explanation for the determined difference.

The accuracy of snowfall measurements is known to be lower than measurements of precipitation (Kuusisto in Mustonen 1986). The FMI snow measurements were melted and measured on a daily basis and thus the evaporation losses were negligible. In the catchments the buckets used for snow collection were emptied once a fortnight or once a month and presumably some evaporation has occurred. The accuracy of the transformation of snow depth measurements into water content in snow cover is difficult to estimate. However, the two different approaches gave rather similar results.

Despite the problems with snowfall measurement, it can be concluded that the estimation of the water input can be taken as one of the most accurate components of the water balance estimation and thus the outcome of water budgets based on other measurements can be compared with the results of precipitation measurements.

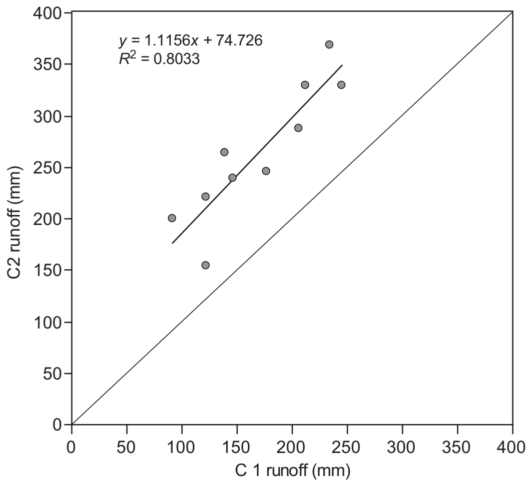

Fig. 14. Correlation between the runoff in the two catchments.

Throughfall

The proportion of canopy interception of the precipitation was high, on an annual basis more than 30%. Especially the canopy of the stand in C1 was very dense. The measured canopy interception is in the range of earlier measurements (Päivänen 1966, Mustajärvi *et al*. 2008). It is possible that the shape of the throughfall gage is such that it can not detect small amounts of throughfall occuring on days when the evaporation is high due to the attachment of the drops on the surfaces of the gutters. It is important to notice that only the throughfall water is available for plants and soil organisms. Also the formation of runoff depends solely on the amount of throughfall.

In some cases one of the seven throughfall counters was not operating correctly. The most common error was a double reading of the counter, but such occasions could be detected and incorrect data removed from the averages. As compared with that of the circular rain gages with the surface area of 200–500 cm^2, the spatial representativeness of the counters was reasonably good, as the total surface area of the gutters was 2.7 m^2 (= 26 950 cm^2).

The trees growing in C2 were smaller than in C1 from the beginning of the measurement period and this difference in needle biomass increased when the stand of C2 was thinned. It was shown that the tree stand of C1 with higher needle biomass caused both higher canopy interception and higher evapotranspiration which was also followed by lower runoff. The effect on evapotranspiration was larger than the effect on interception.

The interception of ground vegetation was not measured here. Based on the irrigation experiment data of Mälkönen *et al*. (1982), it can be concluded that the amount of interception of ground vegetation during a rain episode can be in the range of 1 mm. When taking into account both canopy and ground vegetation interception this means that most of the water in small summer showers never reaches the soil.

The proportion of the stem flow was in the range of 1% of the annual precipitation. This fits well in the limits given by Johnson and Lehman (2006) varying between 0.07–22 for a variety of tree species and precipitation ranges (600–7100 mm).

Soil water storage

In all years, except for the very dry year 2002, the studied soils reached their field capacity in autumn or in early winter. Because the soil water storage was filled up in the autumn and runoff ceased after the accumulation of snow cover had started, the mid winter situation at the change of the calendar year was found to be also the best time to define the start of the new hydrological year. In almost all years the soil was at the field capacity already before the spring snowmelt began, and it was shown that the amount of water released in snowmelt was in close correlation with the amount and timing of runoff and snowmelt. The soil water storage in C2 remained at a higher level during the growing season than the soil water storage in C1, which can be an outcome of smaller evapotranspiration and canopy interception in C2.

The estimation of the amount of changes in soil water storage is primarily based on the estimation of the area of the catchment and the soil depth. Here the depth of the soil was measured with soil radar, and the signal was calibrated with the measured soil depths of the seven pits and two weirs dug down to the bedrock. This method gives a reasonably reliable estimation of the elevation of the bedrock and the variation in the soil depth of the catchments ($R^2 = 0.77$). The borderline between the catchments and the surrounding areas was selected so that it was always the highest point as compared with the both sides of the borderline so that only the water raining within the borders was flowing by gravitation to the weirs, and no uncontrolled flow out of the catchment or inflow from outside the catchments occurred. The definition of the borders of the catchment area is not necessarily very accurate in a terrain which is undulating only gently and having also flat areas with very small angles. It can be shown by a simple calculation that an error of one meter in defining the actual location of the borderline on each side of the catchment can cause a 10%–20% error in the area of C1 and much larger error in C2. When measured runoff values are transformed into mm, errors in area estimations are transferred to the water balance calculations. In the case of this study, precise determinations of soil elevation (accuracy within 1 cm) were used to determine the relative height of each measured corner point of the grid. There are no statistical means to define the magnitude of the possible error in the size of the basins of the two catchments, but due to the shape of the bedrock the surface area of both catchments together is more precise than the two catchments separately as the borderline in the bedrock between the two catchments is not sharp.

The Ledieu calibration of the TDR signal in wet soil conditions proved to be incorrect, since the amount of water in the deeper soil layers showed higher values than the soil water content at pF 0. The pore space at our site equaled 55% volumetric water content in the B-horizon, where the bulk of the soil volume existed, and 47% average volumetric water content through the whole soil profile. The noise in the TDR signal was high especially in the deepest soil horizons, but the erroneous data was reasonably easy to determine and discard, and the number of correct measurements still could give reliable daily averages. The freezing of the soil water causes periods when the water content measurements are not possible with TDR, and it is important to filter such measurements out of the data. However, this change in the dielectricity of the water during freezing is a very useful property to be used in a precise definition of the phase transition from liquid water to ice or vice versa, because in soil temperatures near 0 °C both phases are possible.

Runoff

The timings of the snowmelt and runoff almost coincided. When the amount of water estimated to be released in snowmelt and the measured runoff were in the same range, this can be regarded as a support that both measurements are giving realistic values. However there seems to be a small difference between the runoff calculated as an average weighed by the surface areas of both catchments (193 mm) and snowmelt (150–160 mm) which could show that the soil water content is over the field capacity in deeper soil horizons or the estimation of the amount of water released in the snowmelt is not exactly correct.

The 95 mm difference in the average runoff from C1 and C2 could be interpreted to show the difference in the evapotranspiration between the catchments. This should be the case if the estimations of the catchment areas are correct, which assumption is supported by the fact that the value of the correlation coefficient between the measured runoff of both catchments is near unity (Fig. 14). When the changes in soil water storages of these two catchments are compared, a difference in a similar range can be found.

If the exceptionally dry year 2002 is omitted, the average difference in the runoff between the catchments in the preceding four years was 92 mm, whereas it was 112 mm in the four succeeding years following the thinning. The increase in the difference in the runoff after thinning supports the assumption that higher needle biomass at C1 is causing these differences in the runoff between C1 and C2.

The average runoff of five Swedish sites calculated with SOIL model (Gärdenäs and Jansson 1995) was 160 or 200 mm, depending on the assumptions made concerning soil texture (sandy silt or sand). The calculated runoff was 175 or 210 mm for the Mora site, which locates at approximately the same latitude as the SMEAR-II station. This is in good agreement with the values of this study, 172 and 267 mm for C1 and C2, respectively.

Evapotranspiration

A recent analysis of a large number of sites showed that the energy balance cannot be closed by eddy covariance measurements, possibly because the estimates of latent heat exchange (and therefore evapotranspiration) may be underestimated (Wilson *et al*. 2001). Accordingly, the bias of several tens of percent can be expected for H_2O exchange. Our results obtained with independent measurements of water balance components support this finding. Also Baldocchi *et al*. (1997) presented that the eddy-covariance measurements of water vapor fluxes in boreal conifer forests give underestimated results. Rannik *et al*. (2006) estimated the bias in CO_2 exchange for an eddy-covariance site to be in the order of 30%–50% of the total net exchange. A similar bias may be expected for H_2O exchange because assumptions and methods are similar for CO_2 and H_2O.

The precipitation, throughfall and runoff measurements can be used to estimate the amount of evapotranspiration. The precipitation (692 mm) is assumed to equal the sum of runoff (196 mm) and evapotranspiration (295 mm), but the difference between these values is 201 mm. The difference between the annual average values of througfall (463 mm) and runoff (196 mm) was 267 mm. This difference represents the amount of water transpired from the canopy and it should equal the difference between EC evapotranspiration (295 mm) and canopy interception (229 mm), but was only 66 mm. Based on these comparisons we could conclude that the EC-measured annual evapotranspiration is presumably an underestimation.

The model calculations of Gärdenäs and Jansson (1995) gave evapotranspiration estimations of five Swedish sites in a range between 320 and 580 mm. For the Mora site, the estimated evapotranspiration was 465 mm. The 15-year average precipitation in Walker Branch was 1368 mm and evapotranspiration 655 mm (Luxmoore and Huff 1989). During the growing season the measured throughfall was 610 mm while the precipitation was 721 mm. During the dormant season the corresponding figures were 695 mm and 807 mm. Ladekarl *et al*. (2005) estimated with model calculations verified with eight-year TDR measurements that the evaporation, transpiration and recharge from a Danish oak stand was 205, 285 and 390 mm, respectively. The 40-year-period average of annual precipitation in St. Arnold lysimeter field in Germany was 792 mm and the evaporation from a coniferous forest was 56% and runoff 28% (Harsch *et al*. 2008. The evapotranspiration of a central European declining oak stand varied between 383 and 594 mm (Vincke *et al*. 2005). In this stand the stand transpiration was lower than the forest floor ET.

The catchments studied are located higher than the surrounding areas. The increase in elevation increases precipitation and this difference can be seen, when the annual average precipitation at the SMEAR-II catchments (691 mm) is compared with the average of three surrounding FMI measurement stations (626 mm) (Table 5). When using an annual average, there was 69 mm more precipitation reaching the catchments of this study, than the watersheds of the Finnish Environment Institute (FEI).

The estimated average evapotranspiration by using the $P - R$ difference in the three watersheds (Näsijärvi, Valkeakoski, Vilppula) was 362 mm (Hyvärinen and Korhonen 2003 and hydrological statistics of FEI available in www.environment.fi). In our study the difference between precipitation (692 mm) and average runoff in C1 and C2 (196 mm) was larger, 496 mm (Table 4). Our estimates of ET by difference method are similar to the long term annual evapotranspiration (between 450 and 500 mm) presented for southern Finland (Vakkilainen in Mustonen 1986). The average (1998–2005) ET estimates measured by Class-A evaporation of Jyväskylä, Jokioinen and Vilppula FEI stations between May and September was 441 mm (Table 5). The average evapotranspiration of EC measurements during the summer months stated above was 255 mm. In lake ecosystems a correction factor of 0.7 is used to transform the Class-A evaporation to the lake evaporation (Winter 1981). For terrestrial ecosystems, coefficients of the same magnitude have been presented to convert the Class-A evaporation into potential evapotranspiration. (Vakkilainen in Mustonen 1986). The ratio between ET(EC) of this study and Class-A

Table 5. Cumulative annual values of the components of the water balance (mm) of the ecosystem studied. P = Precipitation, S = precipitation during days when the air temperature was below zero (an estimate of the amount of snowfall), T = Throughfall, I = Canopy Interception, R = Runoff, Tr = Transpiration, ET = Evapotranspiration (EC). C1 and C2 catchments 1 and 2, R_{C1+C2} sum of the runoff in liters from C1 and C2 weirs 1 and 2 divided by the total area (m²) of C1 + C2. Class-A ET is an average of measurements in Jyväskylä, Jokioinen and Vihti. The FEI measurements (values in intalics) are averages of three measurement watersheds (Näsijärvi, Valkeakoski, Vilppula) near the SMEAR-II station.

Year	P	R_{C1+C2}	ET				$P - R_{C1+C2}$ – ET(EC)	Class-A ET	FEI		
			EC	$P - R_{C1+C2}$	$P - R_{C1}$	$P - R_{C2}$			P	R	ET
1998	825	268	235	557	591	457	289	*349*	*684*	*309*	*374*
1999	676	195	233	481	499	431	232	*495*	*602*	*239*	*363*
2000	730	267	270	463	485	401	173	*436*	*663*	*282*	*380*
2001	752	227	356	525	547	464	149	*437*	*675*	*339*	*337*
2002	535	130	340	405	413	381	57	*480*	*530*	*243*	*286*
2003	645	119	320	526	553	445	179	*434*	*601*	*157*	*444*
2004	718	171	294	547	579	454	223	*377*	*633*	*253*	*380*
2005	698	147	–	551	576	477	–	*432*	*641*	*269*	*372*
2006	644	242	313	402	432	314	60		*572*	*210*	*362*
Average	692	196	295	496	519	425	170	*441*	*622*	*256*	*367*

evaporation measured by FEI was 0.58. Based on these comparisons it seems that in our study the evapotranspiration measured using the difference method ($P - R$) gives an overestimation, and evapotranspiration measured with the EC method gives an underestimation.

The discrepancy between the ET(EC) and the $P - R$-method could be explained at least partially by the representativeness of the footprint area of the EC as compared with that of the catchments. However, the average estimated total foliage biomass in the footprint was 5500–6000 kg ha^{-1}, corresponding the needle biomass in C1.

If the average eddy covariance value (295 mm) was correct, it would mean that annually the runoff measurements would include a bias of 178 mm. If the difference of this magnitude in the runoff is transformed into the units of catchment area, the combined area of catchments C1 and C2 should be 656 m^{-2} instead of 1190 m^{-2}. The error of this magnitude in the estimation of the catchment area is not probable.

Conclusions

Our study indicates that the independent measurements of water balance components could be carried out successfully and although the values include some biases they reveal the hydrological properties of the studied boreal Scots pine ecosystem. In all years, the amount of precipitation exceeded the amount of evapotranspiration and each spring runoff occurred. The amount of spring runoff was similar to the amount of water stored in the snowpack. In almost all years the soil profile reached its field capacity during the autumn. The results of the evapotranspiration measurements conducted with the eddy covariance method probably were underestimations, and the evapotranspiration determined from the difference between precipitation and runoff seemed to give overestimations.

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